CHAPTER 27

INFILTRATION AND SOIL MOISTURE PROCESSES

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Infiltration is the process of water entry from surface sources such as rainfall, snowmelt, or irrigation into the soil. The infiltration process is a component in the overall unsaturated *redistribution* process (Fig. 1)[1] that results in *soil moisture* availability for use by vegetation transpiration, exfiltration (or evaporation) processes, chemical transport, and groundwater recharge. Soil moisture, in turn, controls the partitioning of subsequent precipitation into infiltration and runoff, and the partitioning of available energy between sensible and latent heat flux.

Because of the importance of soil moisture on multiple processes, its definition can be elusive[2]; however, it is most often described as moisture in the unsaturated surface layers (first 1 to 2 m) of soil that can interact with the atmosphere through evapotranspiration and precipitation.[3]

1 CONTROLS ON INFILTRATION AND SOIL MOISTURE

To characterize soil moisture and infiltration, the physical controls on these processes must be considered. The primary soil controls will be considered in this chapter; however, other factors such as soil chemistry, thickness, soil layering or horizons, and preferential flow paths, as well as vegetation cover, tillage, roughness, topography, temperature, and rainfall intensity also exert important controls.[4]

A soil's particle size distribution has a large impact on its hydraulic properties. Soil particles less than 2 mm in diameter are divided into three texture groups (sand, silt, and clay) that help to classify broad soil types and soil water responses (Fig. 2).[5] The type of clay and the coarse material over 2 mm in diameter can also have a

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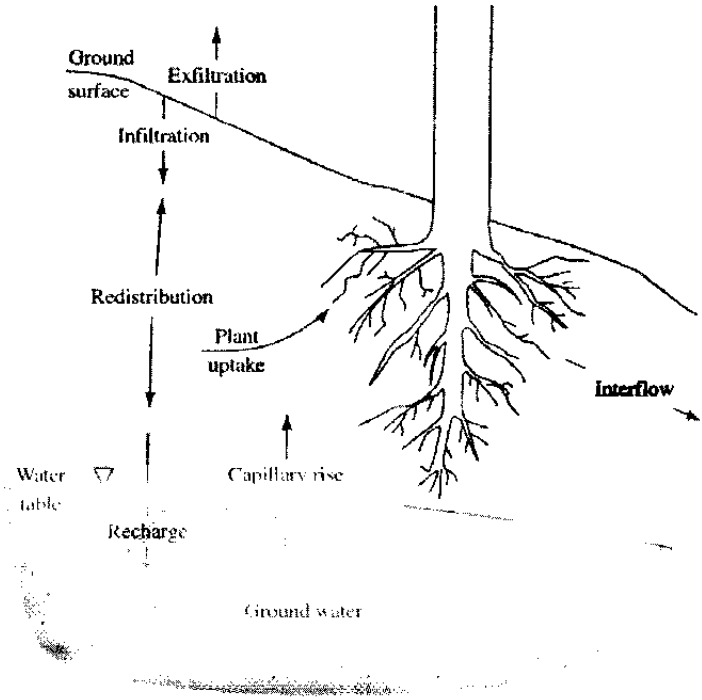

Figure 1 Unsaturated zone definition and active processes.[1]

significant impact on soil water properties. An overview of methods for determining particle size properties is given by Gee and Bauder.[6]

Bulk density, ρ_b (M/L^3) is the ratio of the weight of dry solids to the bulk volume of the soil, and *porosity*, φ (M^3/M^3), is the total volume occupied by pores per unit volume of soil:

$$\varphi = \frac{V_a + V_w}{V_s} = 1 - \frac{\rho_b}{\rho_m} \tag{1}$$

where V_s (L^3) is the total volume of soil, V_a (L^3) is the volume of air, V_w (L^3) is the volume of water, and ρ_m ($M\,L^{-3}$) is the particle density (normally about $2.65\,g/cm^3$).

The volumetric water content, or soil moisture, θ (L^3L^{-3}) is the ratio of water volume to soil volume:

$$\theta = \frac{V_w}{V_s} = \frac{W_w}{W_d}\frac{\rho_b}{\rho_w} \tag{2}$$

where W_w (M) is the weight of water, W_d (M) is the weight of dry soil, and ρ_w (M/L^3) is the density of water. Soil moisture can vary in both time and space, with a

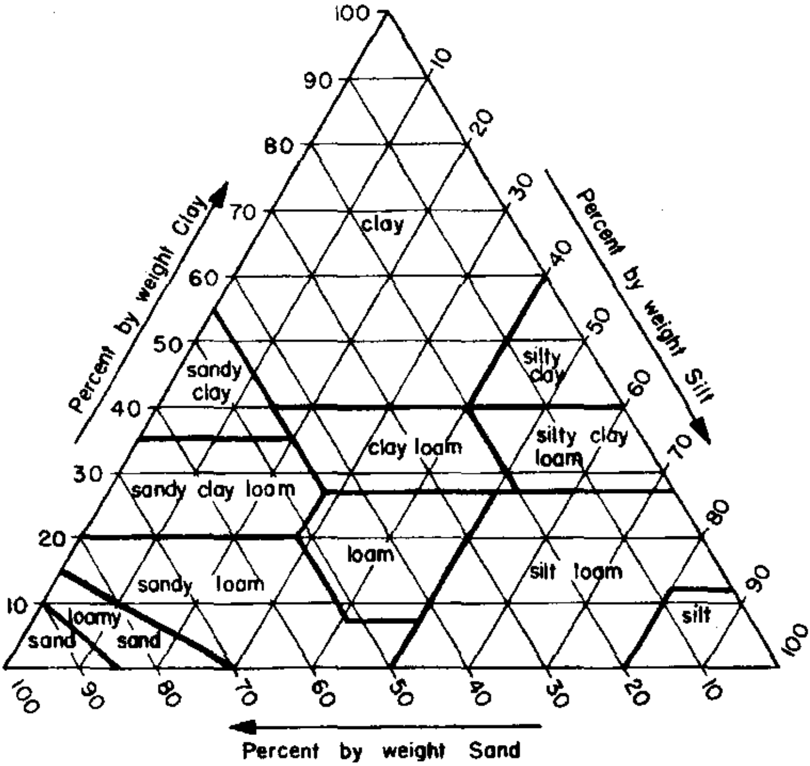

Figure 2 Soil textural triangle describing the relationship between texture and particle size distribution.[5]

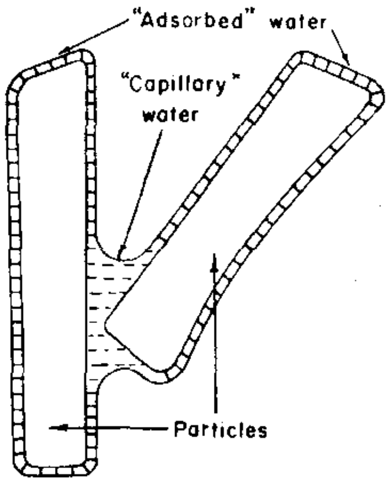

Figure 3 Capillarity and adsorption combine to produce suction.[7]

theoretical range from 0 to φ, but for natural soils the range is significantly reduced due to isolated pore space and tightly held or "adsorbed" water (Fig. 3).[7] If a soil is saturated, then allowed to drain until the remaining water held by surface tension is in equilibrium with gravitational forces, it is at *field capacity*, θ_f. Vegetation can remove water from the soil until the *permanent wilting point*, θ_w, is reached. Therefore, the *available water content* for plant use, $\theta_a = \theta_f - \theta_w$. Typical ranges of porosity, field capacity, and wilting point for different soils are given in Fig. 4.[8]

In unsaturated soils, water is held in the soil against gravity by surface tension (Fig. 3). This tension, suction, or *matric potential*, ψ (L), increases as the radii of curvature of the meniscus or water content decreases (Fig. 5).[9] Matric potential is expressed in reference to atmospheric pressure, so for saturated soil $\psi = 0$ and for unsaturated soil $\psi < 0$.

The *hydraulic conductivity*, K (L/T), is a measure of the ability of the soil to transmit water that varies nonlinearly over a large range depending on both soil

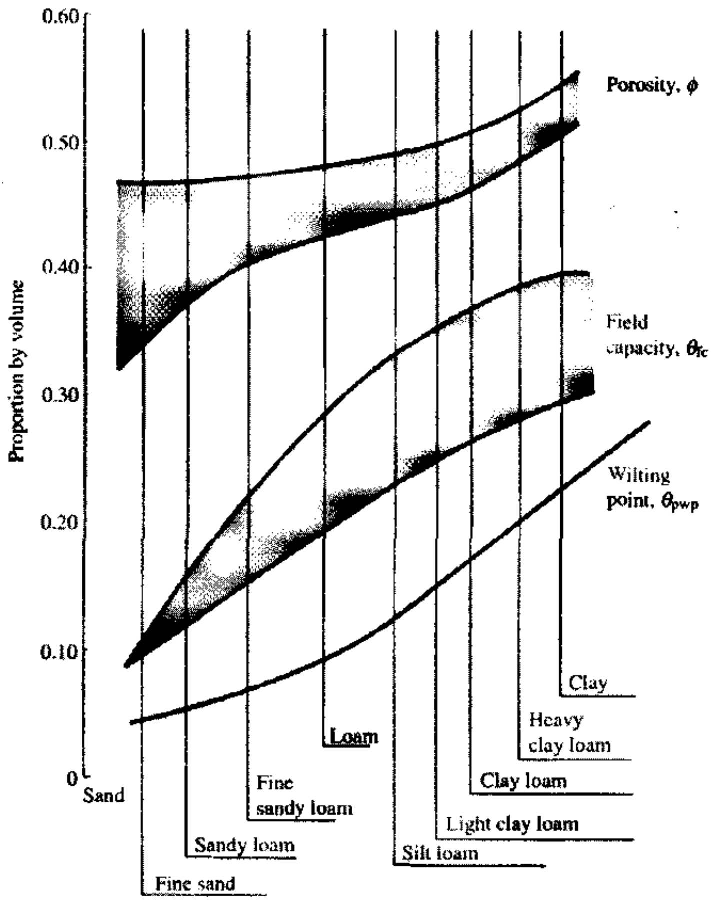

Figure 4 Water holding properties of various soils.[8]

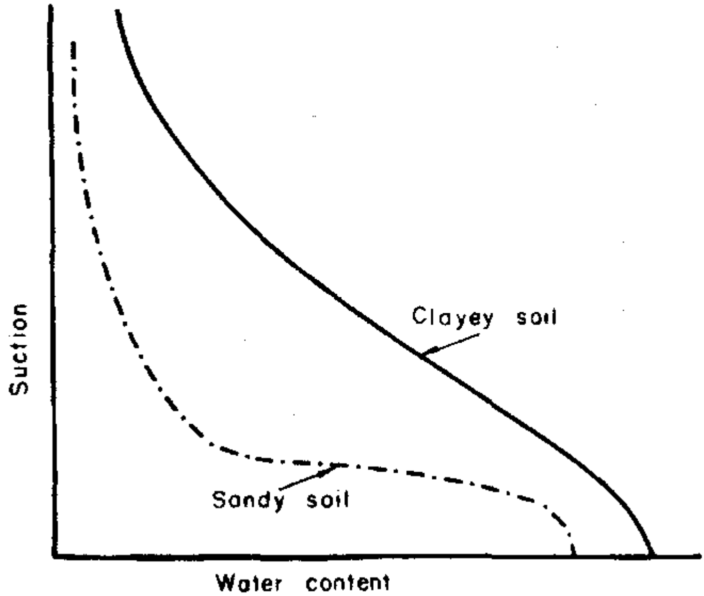

Figure 5 Effect of texture on water retention characteristics.[9]

properties and water content (Fig. 6).[10] Many laboratory and field hydraulic conductivity measurement methods exist for use with various soils; see Bouwer and Jackson[11] or Green et al.[12] for details.

Soil water content can significantly impact infiltration by (1) increasing the hydraulic conductivity, which increases infiltration, and (2) reducing the surface tension that draws moisture into the soil, which reduces infiltration. The net effect of these impacts depends on the water content itself, the water input rate, and duration and the distribution of hydraulic conductivity.

The *water retention characteristic* describes a soil's ability to store and release water and is defined by the relationship between soil moisture and the matric potential (Fig. 5). This is a power function relationship that has been described by Brooks and Corey[13] and Van Genuchten,[14] among others. The water tension characteristic is usually measured in air pressure chambers where the water content of a soil sample can be monitored over a wide pressure range.[15]

The water retention relationship may actually change between drying and wetting due to the entrapment of air in soil pores (Fig. 7).[16] For practical applications, this effect, called *hysteresis*, is usually neglected.[17]

2 PRINCIPLES OF SOIL WATER MOVEMENT

Through experiments on saturated water flow through sand beds, Darcy[18] found that the rate of flow, Q (L^3/T), through a cross-sectional area A (L^2), is directly propor-

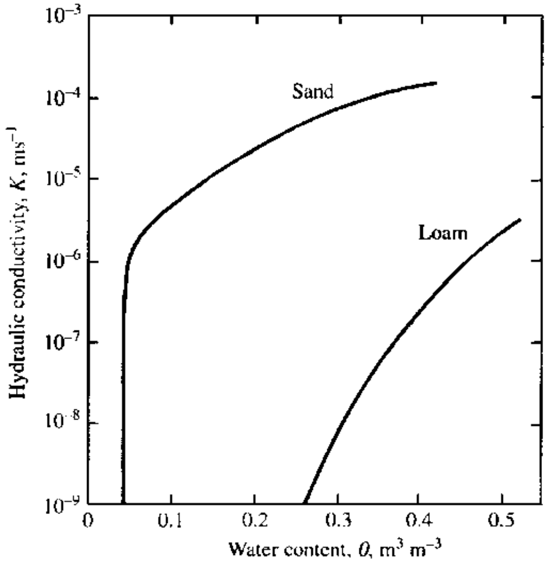

Figure 6 Effect of texture and soil moisture on hydraulic conductivity.[10]

tional to head loss (e.g., water elevation difference), ΔH (L), and inversely to the flow path length, Δl (L):

$$Q = KA\frac{\Delta H}{\Delta l} \tag{3}$$

Combining *Darcy's law* with the law of conservation of mass results in a description of unsaturated flow called *Richards equation*[19]:

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z}\left(\frac{K}{C}\frac{\partial \theta}{\partial z}\right) - \frac{\partial K}{\partial z} \tag{4}$$

where $C = -\partial\theta/\partial\psi$ is the water content change in a unit soil volume per unit matric potential, ψ change. The Richards equation is the basis for most simulations of infiltration and redistribution of water in unsaturated soil. Using some approximations, analytical solutions of the Richards equation are available[20,21] that show good agreement with observations.[22] The Richards equation is based on saturated flow theory, and does not account for all of the processes active in natural systems, so it may not always perform well.[23]

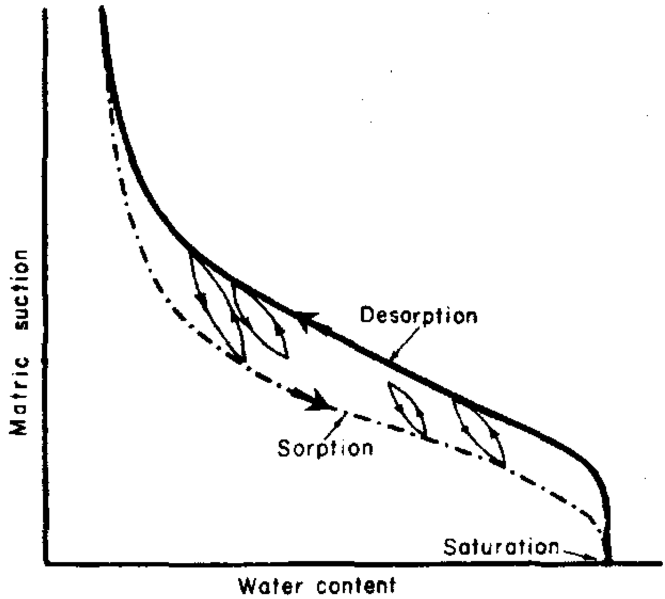

Figure 7 Changes in water retention characteristics between sorption and desorption.[16]

3 INFILTRATION ESTIMATION

Some basic principles that govern the movement of water into the soil can be used to predict infiltration. The *infiltration capacity*, f (L), is the maximum rate that a soil in a given condition can absorb water and generally decreases as soil moisture increases. If the *rainfall rate* is less than the infiltration capacity, then infiltration proceeds at the capacity rate. However, if the rainfall rate exceeds the infiltration capacity, then infiltration proceeds at the capacity rate, and the excess rainfall ponds on the surface or runs off. As the time from the onset of rainfall increases, infiltration rates decrease due to soil moisture increases, raindrop impact, and the clogging of soil pores, until a steady-state infiltration rate is reached (Fig. 8).[24] Existing infiltration models use empirical, approximate, or physical approaches to predict infiltration.[25]

Empirical. Empirical infiltration models generally utilize a mathematical function whose shape as a function of time, t, matches observations and then attempts a physical explanation of the process.

Kostiakov[26] proposed the simple infiltration rate, f (L/T) model:

$$f = \alpha\gamma t^{\alpha-1} \tag{5}$$

where α and γ are constants that have no particular meaning and must be evaluated by fitting the model to experimental data.

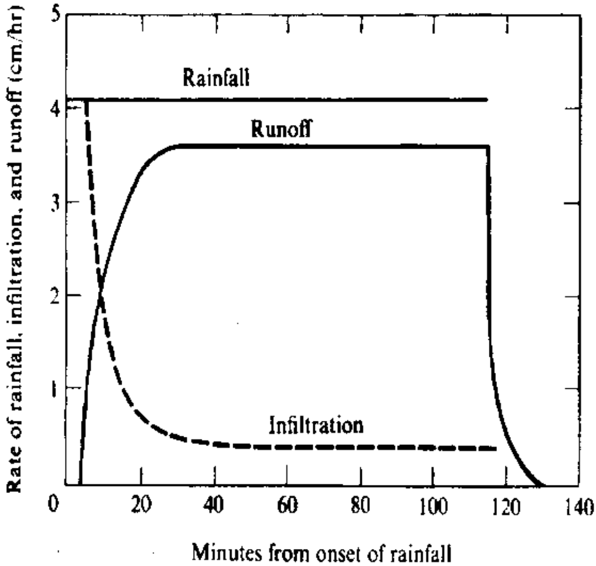

Figure 8 Idealized relationship between rainfall, infiltration, and runoff rates.[24]

Horton's[27] infiltration model has been widely used in hydrologic simulation. It relates infiltration capacity to initial infiltration rate, and f_0, the constant infiltration rate at large times, f_c:

$$f = f_c + (f_0 - f_c)e^{-\beta t} \tag{6}$$

where β is a soil parameter describing the rate of decrease of infiltration.

Approximate. Analysis approximations to the Richards equation are possible if several simplifying assumptions are made. Most approximate infiltration models treat the soil as a semi-infinite medium, with the soil saturating above a wetting front.

Green and Ampt[28] assumed in a soil with constant hydraulic properties, the matric potential at the moving wetting front is constant, leading to a discontinuous change in soil moisture at the wetting front:

$$f = K\left[1 + \frac{(\varphi - \theta_i)S_f}{F}\right] \tag{7}$$

where S_f (L) is the effective suction at the wetting front, θ_i is the initial water content, and F (L) is the accumulated infiltration.

Phillip[29] proposed that the first two terms in a series of powers of $t^{1/2}$ could be used to approximate infiltration:

$$f = \tfrac{1}{2}St^{1/2} + A \tag{8}$$

where S is a parameter called sorptivity, t is time from ponding, and A is a constant that depends on soil properties. In this model, the infiltration rate approaches a constant equal to the hydraulic conductivity at the surface water content, and the wetting front advances without changing its shape and approaches a constant velocity.

Physical. Recent advances in numerical methods and computing has facilitated the practical application of the Richards equation to realistic flow problems. Such packages can simulate water infiltration and redistribution using the Richards equation and including precipitation, runoff, drainage, evaporation, and transpiration processes.[30]

4 INFILTRATION MEASUREMENT

Infiltration rates can be measured at a point using a variety of methods described here, each appropriate for certain conditions. However, because of the large temporal and spatial variability of infiltration processes, catchment average infiltration rates may be desired, which can be obtained through the water balance analysis of rainfall–runoff observations.[31]

Ring Infiltrometer. This simple method is most appropriate for flood irrigation or pond seepage infiltration. A cylindrical metal ring is sealed at the surface and flooded. Intake measurements are recorded until steady-state conditions are reached.[32] If the effects of lateral flow are significant, then a double-ring infiltrometer can be used. Due to ponding conditions within the ring, observed infiltration rates are often higher than under natural conditions.[33]

Sprinkler Infiltrometer. This method is appropriate for quantifying infiltration from rainfall. Artificial rainfall simulators are used to deliver a specified rainfall rate to a well-defined plot. Runoff from the plot is measured, allowing computation of the infiltration rate.[34,35]

Tension Infiltrometer. The tension or disk infiltrometer employs a soil contact plate and a water column that is used to control the matric potential of the infiltrating water. By varying the tension, the effect of different size macropores can be determined.[36,37]

Furrow Infiltrometer. This method is useful if information on infiltration of flowing water in irrigation furrows is desired. Either the water added to a small section of blocked off furrow to maintain a constant depth or the inflow–outflow of a furrow segment can be monitored to determine the infiltration characteristics of the system.[38]

5 SOIL MOISTURE MEASUREMENT

Soil water content can be determined directly using gravimetric techniques or indirectly by inferring it from a property of the soil.[39,40]

Gravimetric. The oven-drying soil moisture measurement technique is the standard for calibration of all other methods but is time consuming and destructive. The method involves obtaining a wet soil sample weight, drying the sample at 105°C for 24 h, then obtaining the dry sample weight [see Eq. (2)].

Neutron Thermalization. High-energy neutrons are emitted by a radioactive source into the soil and are preferentially slowed by hydrogen atoms. The number of slow neutrons returning to the detector are a measure of soil moisture.

Gamma Attenuation. The attenuation in soil of gamma rays emitted from caesium-137 is directly related to soil density. If the soil's bulk density is assumed to be constant, then changes in attenuation reflect changes in soil moisture.[41]

Time-Domain Reflectometry (TDR). TDR measures the soil's dielectric constant, which is directly related to soil moisture, by measuring the transmit time of a voltage pulse applied to a soil probe.

Tensiometric Techniques. This method measures the capillary or moisture potential through a liquid-filled porous cup connected to a vacuum gage. Conversion to soil moisture requires knowledge of the water retention characteristic.

Resistance. The electrical resistance or conductivity of a porous block (nylon, fiberglass, or gypsum) imbedded in the soil depends primarily on the water content of the block. However, because of salinity and temperature sensitivity, measurements of these sensors are of limited accuracy.[42]

Heat Dissipation. Changes in the thermal conductivity of a porous block imbedded in the soil depend primarily on the water content of the block. The dissipation of a heat pulse applied to the block can be monitored using thermistors, then the soil water content can be determined from calibration information.

Remote Sensing. Soil moisture can be remotely sensed with just about any frequency where there is little atmospheric absorption.[43] But, it is generally accepted that long wavelength, passive microwave sensors have the best chance of obtaining soil moisture measurements that contain little error introduced by vegetation and roughness and offer great potential to remotely sense soil moisture content with depth due to differential microwave absorption with varying dielectric constant.[44]

6 SPATIAL AND TEMPORAL VARIABILITY

Natural soils exhibit considerable spatial heterogeneity in both the horizontal and vertical directions, and at all distance scales from the pore to the continent, to a degree that it is difficult to capture this variability in routine measurements.[45,46] This large variation in soil properties, infiltration, and soil moisture over relatively small areas makes it difficult to transfer the understanding of processes developed at a point to catchment scales. Many hydrological models assume that a single spatially representative average soil property can be used to characterize catchment (or even larger) scale processes. It is clear from the nonlinear character of soil water processes [Eq. 94)] that catchment average infiltration cannot be computed based on catchment average soil properties. It is also clear that the physical meaning of a soil property, say porosity, is relative to the volume over which it is averaged.[47] However, there is a need to understand and reduce this complexity for the purposes of prediction and management. Several approaches, including dividing the catchment into hydrologically similar subareas,[48] various statistical approaches,[49] and scaling and similarity theory[50,51] have made headway toward an understanding of infiltration and soil moisture spatial variability, but are not being widely used in practical applications.

One of the most important recent findings in this regard is the scale invariance of soil water behavior. If a heterogenous field is the union of homogenous spatial domains, each with associated characteristic length scales, then heterogeneity simplifies into the spatial variability of these length scales, while the functional relationships that describe soil water movement (i.e., the Richards equation) remain uniform across spatial scales.[52] This new understanding of the underlying symmetry of the Richards equation may help to facilitate a workable scale invariant analytical soil water dynamical model.

Finally, there is a continuing need for the observation of soil properties, soil moisture, and infiltration processes at multiple scales to facilitate understanding and prediction of these complex and socially significant processes. It is likely that remote sensing of soil moisture and other land surface factors will be instrumental in this respect.

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